# Climatic impact of the long-lasting 1783 Laki eruption: Inapplicability of mass-independent sulfur isotopic composition measurements

Anja Schmidt,[1] Thorvaldur Thordarson,[2] Luke D. Oman,[3] Alan Robock,[4] and Stephen Self[5]

Received 3 July 2012; revised 21 September 2012; accepted 23 October 2012; published 13 December 2012.

[1] The long-lasting 1783–1784 CE Laki flood lava eruption in Iceland released around 120 Tg of sulfur dioxide into the upper troposphere/lower stratosphere. Northern Hemisphere temperature proxy records of the 1780s indicate below-average temperatures for up to three years following the eruption. The very warm summer of 1783 in Europe, which was followed by a very cold winter, may have been caused by the eruption, but the mechanisms are not yet well understood. Some studies attributed the cold winter 1783–1784 to natural variability of climate. However, our climate model simulations show that the Laki radiative effects lasted long enough to contribute to the winter cooling. We suggest that sulfur isotopic composition measurements obtained using samples from Greenland ice cores do not provide evidence of either a short-lived volcanic aerosol cloud or a short-lived climatic impact of the Laki eruption. In fact, the applicability of mass-independent sulfur isotopic composition measurements for interpreting the climatic impact of any high-latitude eruption remains yet to be demonstrated.

**Citation:** Schmidt, A., T. Thordarson, L. D. Oman, A. Robock, and S. Self (2012), Climatic impact of the long-lasting 1783 Laki eruption: Inapplicability of mass-independent sulfur isotopic composition measurements, *J. Geophys. Res.*, *117*, D23116, doi:10.1029/2012JD018414.

## 1. Introduction

[2] The 1783–1784 CE Laki eruption commenced on 8 June 1783 and is the best-documented example of a long-lasting flood lava eruption in Iceland. Over the course of eight months, the eruption injected around 120 Tg of sulfur dioxide ($SO_2$) into the upper troposphere and lower stratosphere above Iceland [*Thordarson and Self*, 2003]. Generally, the effects of powerful but relatively short-lived explosive volcanic eruptions releasing $SO_2$ and other gases into the stratosphere on radiative forcing of climate and atmospheric circulation are fairly well understood [e.g., *Robock*, 2000]. Nevertheless, there is now an emerging debate as to the lifetime of the volcanic aerosol, the length of the radiative forcing and the magnitude of the climate impact of the long-lasting 1783–1784 CE Laki eruption [*D'Arrigo et al.*, 2011; *Lanciki et al.*, 2012].

[3] In January 2012, Laki-type eruptions were added to the UK National Risk Register for Civil Emergencies as one of the highest priority risks with potentially "widespread impacts on health, agriculture and transport" [*UK National Risk Register*, 2012]. Previous studies have assessed the health hazard such an event could pose for the population in contemporary Europe [*Schmidt et al.*, 2011], and the atmospheric and climatic perturbations created by the eruption in 1783–1784 [*Highwood and Stevenson*, 2003; *Stevenson et al.*, 2003; *Thordarson and Self*, 2003; *Chenet et al.*, 2005; *Oman et al.*, 2006a, 2006b; *Schmidt et al.*, 2010].

[4] Even though the occurrence of a large-magnitude Laki-type event is of a much lower probability (one event every 200 to 500 years) than that of a typical, much-smaller-magnitude explosive Icelandic volcanic eruption (one event every five years on average) [*Thordarson and Larsen*, 2007], it is important to fully understand the length of the climate impact of a Laki-type event for hazard mitigation and emergency planning. We therefore review and further analyze data from existing studies [*Oman et al.*, 2006a, 2006b] to argue that the suggestion of a short-lived (i.e., less than six months) climate impact of the 1783–1784 CE Laki eruption [*D'Arrigo et al.*, 2011; *Lanciki et al.*, 2012] is unsupported. We discuss arguments for and against a considerable atmospheric/climatic perturbation with a focus on the study by *Lanciki et al.* [2012], who used mass-independent sulfur isotopic composition measurements of the Laki sulfates to interpret Laki's climatic impact. We also suggest that the estimate of the stratospheric

[1]School of Earth and Environment, University of Leeds, Leeds, UK.
[2]School of GeoSciences, Grant Institute, University of Edinburgh, UK.
[3]NASA Goddard Space Flight Center, Greenbelt, Maryland, USA.
[4]Department of Environmental Sciences, Rutgers University, New Brunswick, New Jersey, USA.
[5]Department of Earth and Environmental Sciences, The Open University, Milton Keynes, UK.

Corresponding author: A. Schmidt, School of Earth and Environment, University of Leeds, Leeds LS2 9JT, UK. (a.schmidt@leeds.ac.uk)


0148-0227/12/2012JD018414

component from Laki by *Crowley and Unterman* [2012] is too simplistic and not supported by deposition fluxes simulated using comprehensive general circulation models [*Oman et al.*, 2006a; *Gao et al.*, 2007].

## 2. Arguments for Inferring a Short-Lived Climate Impact

[5] *Lanciki et al.* [2012] presented mass-independent sulfur isotopic composition ($\Delta^{33}S$) measurements of a volcanic sulfate ($SO_4$) peak identified at ~20 m depth in a 79-m ice core from the Summit region of the Greenland ice cap. This $SO_4$ peak has been linked to the Laki eruption and four samples from the "Laki" layer were collected for analysis. Three out of four samples yielded measurable results with two samples showing no significant $\Delta^{33}S$ anomaly ($\Delta^{33}S \approx 0$ ‰) and one sample showing a $\Delta^{33}S$ anomaly just outside the analytical uncertainty range ($\Delta^{33}S = -0.13$ ‰).

[6] Previous studies provided evidence that volcanic $SO_4$ aerosol formed during powerful explosive eruptions in the tropics and recorded in Antarctic ice cores possesses a significant $\Delta^{33}S$ anomaly [*Savarino et al.*, 2003; *Baroni et al.*, 2007, 2008]. *Lanciki et al.* [2012] showed that the $SO_4$ aerosol extracted from one ice core in Greenland linked to the tropical eruption of Tambora (1815) and two eruptions (1806–1809 CE and 1256–59 CE) from unidentified volcanoes (hereafter "Unknown") also possess significant $\Delta^{33}S$ anomalies.

[7] Tambora was a large explosive eruption in the tropics injecting volcanic material high into the stratosphere, and the Unknown events are assumed to have also been large explosive eruptions [e.g., *Stothers*, 1984; *Oppenheimer*, 2003; *Self et al.*, 2004; *Gao et al.*, 2008; *Thordarson et al.*, 2009] (Figure 1a). For such eruptions, the $\Delta^{33}S$ signature measured in the volcanic $SO_4$ has been suggested as a means to assess whether the photochemical conversion of $SO_2$ to $SO_4$ aerosol took place in the stratosphere at altitudes >20 km or whether it was confined to altitudes below that (i.e., the lower stratosphere or upper troposphere) [e.g., *Savarino et al.*, 2003]. Therefore, the presence of $\Delta^{33}S$ anomalies might provide some indication of the initial injection height of the sulfur-bearing volcanic plumes produced by powerful explosive eruptions in the tropics, and the eruption's subsequent climatic impact [*Savarino et al.*, 2003; *Baroni et al.*, 2007, 2008; *Lanciki et al.*, 2012]. However, no previous study has shown whether sulfur isotopic composition measurements and their interpretation are applicable to volcanic eruptions at high latitudes in the same manner as for powerful explosive eruptions in the tropics.

[8] Depending on the chemical process and reaction conditions, fractionation of sulfur isotopes results in sulfur isotope ratios ($\delta^{32}S$, $\delta^{33}S$, $\delta^{34}S$, $\delta^{35}S$) measured in the reaction product either being proportional to the relative difference in mass between the lighter and the heavier isotope (i.e., mass-dependent with, for example, $\delta^{33}S \approx 0.515\ \delta^{34}S$) or deviating from this mass-dependent relationship (i.e., mass-independent or anomalous, for example, $\Delta^{33}S = \delta^{33}S - 1000\ [(1 + \delta^{34}S/1000)^{0.515} - 1]$) [e.g., *Farquhar et al.*, 2001]. At altitudes below 17 km, the gas-phase oxidation of $SO_2$ into sulfuric acid by the hydroxyl radical (OH) is mass-dependent. Hence, $SO_4$ aerosol particles formed in the troposphere are expected to carry no anomalous $\Delta^{33}S$ signals (Figure 1b). In the stratosphere the gas-phase oxidation of $SO_2$ into sulfuric acid proceeds via the same reactions as in the troposphere (i.e., reactions (R1) to (R4) in Figure 1b) but photochemical reactions driven by ultraviolet (UV) radiation, such as the photolysis of either $SO_2$ or $SO_3$, result in a mass-independent fractionation of sulfur isotopes, the signature of which could then be transferred as a marked $\Delta^{33}S$ anomaly to the $SO_4$ aerosol product (Figure 1b) [*Farquhar et al.*, 2001; *Savarino et al.*, 2003; *Pavlov et al.*, 2005]. $SO_2$ photolysis has been shown to take place at UV wavelengths shorter than 220 nm [e.g., *Farquhar et al.*, 2001] and is therefore largely confined to stratospheric altitudes above 20 km. $SO_3$ photolysis requires UV wavelengths between 195 nm and 330 nm, which can be present in the troposphere; however, $SO_3$ photolysis is most efficient at mid-stratospheric altitudes between 25 km and 35 km. [e.g., *Burkholder and McKeen*, 1997]. *Savarino et al.* [2003] proposed a third reactive mechanism by which the $SO_4$ aerosol product could incorporate a $\Delta^{33}S$ anomaly based on a "photo-oxidation" reaction suggested by *Chung et al.* [1975]. *Pavlov et al.* [2005] used a 2-D model to simulate the $\Delta^{33}S$ signature of the volcanic $SO_4$ aerosol formed during the 1991 Mt. Pinatubo eruption (15.13°N, 120.35°E) and concluded that, in the present-day atmosphere, $SO_2$ photolysis could not produce any measured $\Delta^{33}S$ anomaly in ice cores, but that $SO_3$ photolysis could. Overall, there is still debate as to which photochemical reaction pathways are responsible for inducing $\Delta^{33}S$ anomalies in sulfates in the present-day atmosphere. Further research is also required to quantify the chemical and physical processes involved on different temporal and spatial scales.

[9] Notwithstanding the above, based on their $\Delta^{33}S$ measurements with $\Delta^{33}S \approx 0$ ‰ for two out of the four Laki $SO_4$ samples, *Lanciki et al.* [2012] concluded that:

[10] i) "… the Laki eruption injected no significant amount of $SO_2$ into the stratosphere above the maximum ozone altitudes" (i.e., >20 km).

[11] ii) "The deposition pattern of the Laki sulfate in Greenland ice cores indicates a short (<6 months) residence time for the bulk of the Laki aerosols, consistent with a tropospheric-only scenario of Laki aerosols." and

**Figure 1.** Schematics outlining dispersal directions, deposition and chemical processes relevant for powerful volcanic eruptions in the tropics (Pinatubo) and long-lasting high-latitude flood lava eruptions (Laki). (a) The main dispersal pathways in the stratosphere and removal of the aerosol in midlatitude storm belts and regions of tropopause folds for the Mt. Pinatubo eruption and the Laki eruption. (b) Schematic representations of the mass-dependent photochemical conversion of sulfur dioxide ($SO_2$) to sulfate aerosol ($SO_4$), with the $SO_4$ aerosol not possessing a $\Delta^{33}S$ anomaly (black circles) for (I) the 1783–1784 CE Laki eruption supporting eruption columns that reached the upper troposphere (UT) and lower stratosphere (LS) above Iceland. Panels (II) and (III) are representative of more powerful explosive volcanic eruptions that yield eruption columns in excess of 20 km altitude, with panel (II) showing that the mass-independent photolysis of sulfur trioxide ($SO_3$) at altitudes above 25 km can result in a $\Delta^{33}S$ anomaly in the $SO_4$ aerosol product (gray circles) [cf. *Pavlov et al.*, 2005]; and panel (III) shows an alternative reaction pathway via the photolysis of $SO_2$ [*Farquhar et al.*, 2001] (gray circles in panel III), that has been challenged by *Pavlov et al.* [2005] (black circles in panel III).

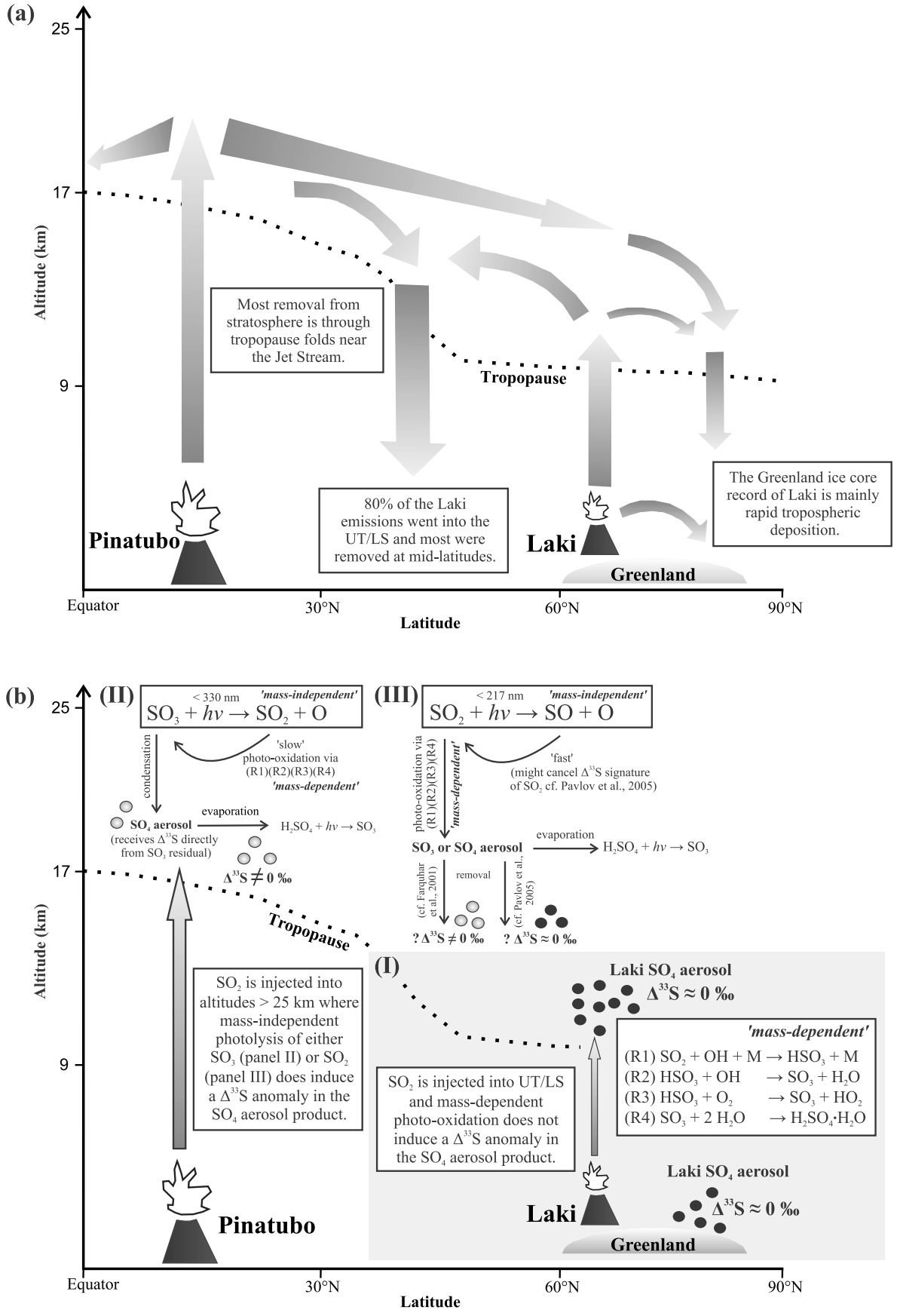
(a)
25
17
9
Altitude (km)
Most removal from stratosphere is through tropopause folds near the Jet Stream.
Tropopause
80% of the Laki emissions went into the UT/LS and most were removed at mid-latitudes.
The Greenland ice core record of Laki is mainly rapid tropospheric deposition.
Pinatubo
Laki
Greenland
Equator
30°N
60°N
90°N
Latitude
(b)
(II)
< 330 nm 'mass-independent'
SO3 + hν → SO2 + O
condensation
'slow' photo-oxidation via (R1)(R2)(R3)(R4) 'mass-dependent'
evaporation
SO4 aerosol
(receives Δ33S directly from SO3 residual)
H2SO4 + hν → SO3
Δ33S ≠ 0 ‰
(III)
< 217 nm 'mass-independent'
SO2 + hν → SO + O
photo-oxidation via (R1)(R2)(R3)(R4) 'mass-dependent'
'fast' (might cancel Δ33S signature of SO2 cf. Pavlov et al., 2005)
SO3 or SO4 aerosol
evaporation
H2SO4 + hν → SO3
removal
(cf. Farquhar et al., 2001)
(cf. Pavlov et al., 2005)
? Δ33S ≠ 0 ‰
? Δ33S ≈ 0 ‰
Tropopause
SO2 is injected into altitudes > 25 km where mass-independent photolysis of either SO3 (panel II) or SO2 (panel III) does induce a Δ33S anomaly in the SO4 aerosol product.
(I)
Laki SO4 aerosol
Δ33S ≈ 0 ‰
'mass-dependent'
(R1) SO2 + OH + M → HSO3 + M
(R2) HSO3 + OH → SO3 + H2O
(R3) HSO3 + O2 → SO3 + HO2
(R4) SO3 + 2 H2O → H2SO4·H2O
SO2 is injected into UT/LS and mass-dependent photo-oxidation does not induce a Δ33S anomaly in the SO4 aerosol product.
Laki SO4 aerosol
Δ33S ≈ 0 ‰
Pinatubo
Laki
Greenland
Equator
30°N
60°N
90°N
Latitude

**Figure 1**

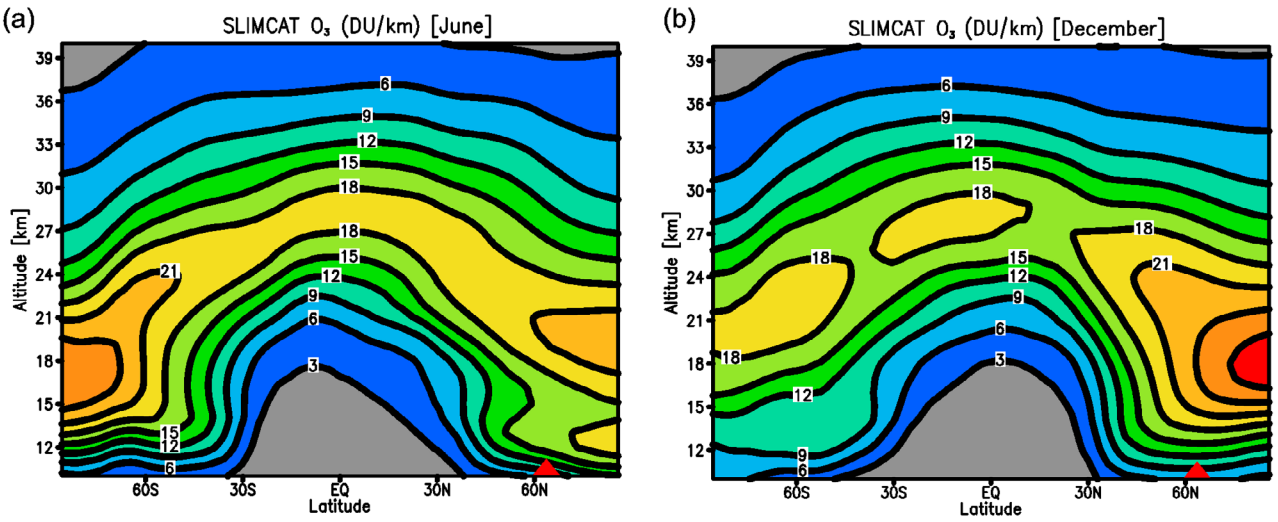


**Figure 2.** Simulated multiyear (1981–2005) zonal mean ozone ($O_3$) concentrations (in DU per km) for (a) June and (b) December using the SLIMCAT chemical transport model. The location of the 1783–1784 CE Laki eruption is indicated by the red triangle.

[12] iii) "Consequently, no significant climatic impact may be expected from the Laki eruption beyond the initial effects and, as suggested by *D'Arrigo et al.* [2011], other factors were probably responsible for the unusual winter of 1783–1784 in the Northern Hemisphere."

[13] We suggest that the second and third of these conclusions are incorrect, that the first one does not imply the second or third one, and that even if the second one were correct, it would not imply the third one.

## 3. Discussion

[14] We review the existing Laki literature, and further analyze data from a previous Laki modeling study [*Oman et al.*, 2006a, 2006b] to discuss whether the apparently short atmospheric residence time of the Laki aerosol together with the apparent lack of any $\Delta^{33}S$ anomalies can be used to deduce a short-lived climatic effect as has been done by *Lanciki et al.* [2012].

[15] To simulate the climatic effect of the Laki eruption, we used the NASA Goddard Institute for Space Studies ModelE general circulation model [*Schmidt et al.*, 2006]. This version of the model has 4° latitude by 5° longitude horizontal resolution with 23 vertical levels up to 80 km. The simulations used a q-flux mixed-layer ocean, which allows the sea surface temperatures to adjust to changing atmospheric fluxes, and a dynamic sea ice model which allows the sea ice to respond to wind.

[16] We conducted ten ensemble members of three years each with varying initial conditions to account for internal natural variability [*Oman et al.*, 2006b]. The control run was based on 30 years of model simulation and all simulations included pre-industrial greenhouse gases and ozone concentrations. The Laki simulations were perturbed with monthly mean volcanic $SO_4$ aerosol concentrations calculated previously [*Oman et al.*, 2006a] with ModelE coupled to an interactive aerosol/chemistry module [*Koch et al.*, 2006], using the Laki $SO_2$ emissions data set of *Thordarson et al.* [1996]. *Oman et al.* [2006a] give a more detailed description of the chemistry module and the relevant $SO_2$ oxidation pathways.

### 3.1. Injection Height of Volcanic Volatiles During Laki

[17] None of the previous work on the Laki eruption suggested that the eruption injected $SO_2$ (or any other material) "above the maximum ozone altitude" or that the eruption resulted in stratospheric loading of sulfur at altitudes required for the mass-independent fractionation of sulfur isotopes to take place. Figure 2 shows simulated multiyear zonal June–mean and December–mean ozone concentrations using the SLIMCAT chemical transport model [*Chipperfield*, 1999, 2006]. At high northern latitudes ozone concentrations are highest in the stratosphere at altitudes between 17 and 21 km, which is well above the suggested eruption column heights of 9 to 13 km reached during the first five months of the Laki eruption that produced ten explosive episodes of sub-Plinian intensity [*Thordarson and Self*, 1993, 2003]. During the summer months the tropopause above Iceland is situated at around 11 km; hence the Laki volatiles were released into the upper troposphere (UT) and lower stratosphere (LS). *Thordarson and Self* [2003] used eruption column models [*Stothers et al.*, 1986; *Thordarson and Self*, 1993; *Woods*, 1993] and an eye-witness record [*Stephensen*, 1783] to derive these column height estimates. Therefore, conclusion (i) by *Lanciki et al.* [2012] corroborates previous Laki studies in terms of the eruption column heights.

### 3.2. Atmospheric Transport, Residence Time and Deposition of the Laki Aerosol

[18] *Lanciki et al.* [2012] concluded that the Laki acidity signal recorded in the Greenland Summit ice core "… favor[s] the scenario of low-altitude injection used in model simulations [*Highwood and Stevenson*, 2003; *Oman et al.*, 2006a]." because "The deposition pattern … indicates a short (<6 months) residence time for the bulk of the Laki aerosols ...."

[19] First, all Laki modeling studies undertaken so far have assumed an $SO_2$ injection height between 9 and 13 km

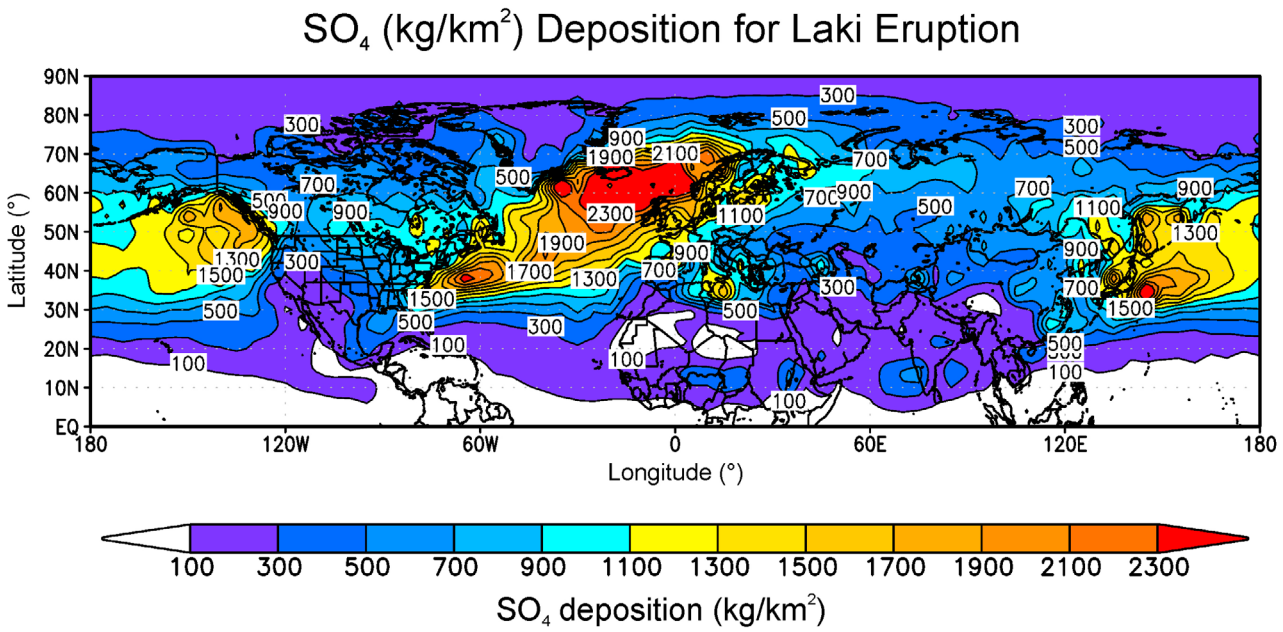


**Figure 3.** Total sulfate ($SO_4$) aerosol deposition (kg/km$^2$) in the year following the onset of the Laki eruption averaged from the three simulations presented in *Oman et al.* [2006a].

[*Highwood and Stevenson*, 2003; *Stevenson et al.*, 2003; *Chenet et al.*, 2005; *Oman et al.*, 2006a, 2006b; *Schmidt et al.*, 2010, 2011]. In addition, *Stevenson et al.* [2003] and *Highwood and Stevenson* [2003] simulated alternative scenarios including a purely tropospheric scenario (referred to as the "low-altitude injection scenario" in *Lanciki et al.* [2012]) in which the $SO_2$ mass was distributed evenly in grid boxes between the surface and around 9 km. Regardless of the employed injection height scenario, all studies mentioned-above indicated the presence of a UT aerosol veil until at least March 1784 (e.g., *Stevenson et al.* [2003, Figure 8] show a comparison of several injection height scenarios using the STOCHEM model).

[20] Second, based on the width of the acidity signal given in Figure 1 of *Lanciki et al.* [2012], the data appear to record more than six months of Laki fallout to the Greenland Summit region. Furthermore, in *Lanciki et al.* [2012], the Laki $SO_4$ peak appears to start in early spring 1783 when it is expected to start in early summer 1783 (because the Laki eruption commenced on 8 June 1783), which might suggest issues with their age-model.

[21] Third, based on seasonal wind direction profiles above Iceland [*Lacasse*, 2001] and dispersal directions for recent explosive eruptions in Iceland [*Petersen et al.*, 2012] one can conclude that most of the UT/LS component of the Laki aerosol veil is expected to have been dispersed eastward. Historical records in Iceland as well as dispersal axes obtained from the Laki tephra fallout on the ground in Iceland and other parts of Northwestern Europe provide further evidence of a net eastward transport [*Thordarson and Self*, 1993; *Thordarson et al.*, 2003]. Analysis of these records, including contemporary synoptic weather maps [*Kington*, 1988], not only indicate that the bulk transport took place at the polar jet stream level but also show that there was no significant westward plume dispersal toward Greenland [*Thordarson and Self*, 1993; *Thordarson et al.*, 2001; *Thordarson and Self*, 2003; *Thordarson et al.*, 2003]. In fact, regions west of the Laki fissure were the areas in Iceland that were affected the least [*Thordarson and Self*, 2003, and references therein]. The contemporary records also describe the occurrence of a lower tropospheric "dry fog" in certain regions of western Greenland – the timing of which matches the first occurrence of a dry fog in Northern Iceland. Therefore, those plumes that reached Greenland were most likely tropospheric, first transported to the east or northeast by southwesterly winds followed by subsidence and deposition onto the Greenland ice cap after circling the Polar region (Figure 1a). For the Laki eruption, it is therefore reasonable to conclude that only a small proportion of the UT/LS aerosol loading was deposited onto the Greenland ice cap. Hence, there is a strong possibility that *Lanciki et al.* [2012] mainly sampled the tropospheric component of the Laki aerosol veil.

[22] *Crowley and Unterman* [2012] suggested that the deposition of Laki $SO_4$ in western Canada can be used to estimate the stratospheric component of the Laki eruption; however, we suggest that climate model simulations make a more comprehensive effort to model the physics of the situation. *Oman et al.* [2006a] used a general circulation model to show that the Greenland component of the Laki $SO_4$ deposition matches the deposition fluxes as recorded in 23 different ice cores. Figure 3 shows the total $SO_4$ deposition in the year following the Laki eruption using the model simulations by *Oman et al.* [2006a]. For the Laki eruption, it becomes evident that the average deposition flux to Greenland is lower than the Northern Hemisphere (NH) average, which is corroborated by *Stevenson et al.* [2003, Figure 9] for all three injection height scenarios used in the STOCHEM model. In fact, the main deposition region for a stratospheric aerosol veil from a high-latitude eruption in the NH is found in midlatitude storm belts and regions of tropopause folds (see also Figure 1a) which is in agreement with other climate modeling studies [*Gao et al.*, 2007;

*Kravitz and Robock*, 2011]. Given the discussion above, we conclude that *Lanciki et al.* [2012] do not categorically show that the Laki aerosol veil cannot have been present at UT/LS altitudes for more than six months due to the close proximity of the eruption and deposition site. In addition, there is no means to infer whether the analyzed samples from Greenland's Summit region are representative of the stratospheric or tropospheric component of the Laki aerosol veil or a mixture of both. We suggest that, due to the proximity of Greenland to the Laki eruption site, $SO_4$ deposition to Greenland ice cores may be an inadequate and ambiguous proxy for drawing conclusions on the residence time of Laki $SO_4$ aerosol, especially that present in the UT/LS. *Baroni et al.* [2007, 2008] noted that for mass-independent sulfur isotope composition analysis, samples from several locations should be analyzed because both the flux of the $SO_4$ deposition to snow and its preservation vary considerably across the polar ice caps. Such findings are reinforced by the fact that there is a large spatial variability in the Laki $SO_4$ deposition fluxes among Greenland ice cores with fluxes ranging from $\sim$80 kg km$^{-2}$ to >300 kg km$^{-2}$ [e.g., *Clausen and Hammer*, 1988; *Mosley-Thompson et al.*, 2003; *Oman et al.*, 2006a; *Gao et al.*, 2007].

### 3.3. Interpretation of $\Delta^{33}$S Measurements for High-Latitude Eruptions

[23] *Baroni et al.* [2008] noted that only high-time-resolution sampling of one volcanic event will give confidence as to whether an eruption was stratospheric in nature (if so, the sign of the $\Delta^{33}$S anomaly changes over the course of the $SO_4$ deposition onto the ice cap). Similarly, *Castleman et al.* [1973, 1974] noted time-dependent changes in $\delta^{34}$S isotope ratios measured in particles sampled in the lower stratosphere following the 1963 Mt. Agung eruption (8.5°S [*Self and Rampino*, 2012]). *Lanciki et al.* [2012] collected four samples across the Laki $SO_4$ peak in order to measure $\delta^{33}$S and $\delta^{34}$S; however, the sample corresponding to the peak $SO_4$ concentration contained too little $SO_4$ mass to enable analysis. It is also noteworthy that the first Laki sample (Sample 1 in *Lanciki et al.* [2012]) actually possesses a $\Delta^{33}$S anomaly of $-0.13$ ‰ that is outside the $\Delta^{33}$S analytical uncertainty range of $\pm 0.08$ ‰. Thus *Lanciki et al.* [2012] could have given more consideration to the following aspects: 1) There are not enough measurements to characterize the Laki deposition over time, and the Greenland ice cap might not be the most suitable sampling location due to the close proximity of eruption site and deposition site. We doubt that there is a sufficient spatial and temporal separation of the isotopic reservoirs when samples from the Greenland ice cap are analyzed for high-latitude eruptions. 2) Using measurements from one core only, when it is known that there can be a high spatial variability in sulfate deposition on ice, further decreases the significance of the signal. 3) Arguing that positive $\Delta^{33}$S anomalies of +0.18 ‰ for the Unknown 1809 event and +0.23 ‰ for Tambora 1815 are "large" and significant anomalies while the $\Delta^{33}$S anomaly of $-0.13$ ‰ for Laki is insignificant (seemingly because of the opposing negative sign) seems ambiguous and not well justified. 4) If $SO_2$ photolysis were the driving process then one would not expect the Laki $SO_4$ to possess any $\Delta^{33}$S signature because the Laki volatiles did not reach altitudes above 17 km where $SO_2$ photolysis takes place. If $SO_3$ photolysis were the driving process then the Laki $SO_4$ could, in theory, carry a $\Delta^{33}$S anomaly. However, inferring the altitude of volatile injection would be very ambiguous given the wide UV wavelength range. 5) Using $\Delta^{33}$S measurements to interpret the climatic impact of any eruption, consideration needs to be given to differences in eruption style, eruption duration and source location. For example, the main phase of the Tambora eruption lasted for about 18 h versus repeated explosive episodes (total of ten) over the course of the first five months for Laki.

[24] We conclude that, so far, neither the process inducing a $\Delta^{33}$S anomaly in the volcanic sulfate product nor the applicability of the measurements to high-northern latitude eruptions has been rigorously demonstrated. *Bindeman et al.* [2007] further discuss the limitations of $\Delta^{33}$S and $\Delta^{34}$S measurements in sulfates leached from volcanic ash deposits for a range of eruptions of different magnitudes and concluded that a lack of a mass-independent sulfur isotope signal does not necessarily rule out that the volcanic volatiles (for eruptions other than Laki) reached stratospheric altitudes. *Bindeman et al.* [2007] and *Martin and Bindeman* [2009] suggested that using oxygen isotope measurements in synergy with sulfur isotopes might be a more adequate means to interpret photolytic mechanisms, aerosol transport and deposition processes as well as the climate impact of volcanic eruptions.

### 3.4. Observed and Modeled Climate Impact Following Laki

[25] *Lanciki et al.* [2012] used the lack of a $\Delta^{33}$S anomaly together with the apparently short residence time of the Laki aerosol as deduced from the Summit ice core to argue that "[…] any Laki impact on NH climate would be minimal and short-lived (i.e., limited to 1783) […]." referring to the purely tropospheric scenario as used in *Highwood and Stevenson* [2003]. However, the results of *Highwood and Stevenson* [2003] indicate that, in the NH, the direct radiative forcing of climate due to the Laki $SO_4$ aerosol lasted at least until March 1784. Even though the magnitude of the radiative forcing varies depending on the injection height scenario (maximum global mean forcing between $-3$ W m$^{-2}$ and $-1.5$ W m$^{-2}$), the forcing is still substantial with the purely tropospheric scenario resulting in a maximum NH mean radiative forcing of $-3$ W m$^{-2}$ [*Highwood and Stevenson*, 2003].

[26] We suggest that *Lanciki et al.* [2012] do not recognize that the climatic impact of the eruption will outlast the radiative forcing anomaly. We highlight this point in Figure 4 showing that the simulated changes in NH surface temperatures lasted well into the latter half of 1784 and clearly outlast the $SO_4$ aerosol optical depth perturbation. *Oman et al.* [2006a] simulated a peak global mean forcing of $-4$ W m$^{-2}$ for August 1783 resulting in a summer surface temperature anomaly of $-1°$ to $-3$°C over large parts of the NH [*Oman et al.*, 2006b]. Averaged over the NH, we find a significant cooling of about $-1$°C during September 1783 (Figure 4a) with a peak aerosol optical depth of about 0.5 one month earlier. *Highwood and Stevenson* [2003] found no statistically significant (at the 95% confidence level) surface temperature anomaly for the purely tropospheric scenario; however, NH temperature anomalies of up to $-0.35$ K were found in the high-altitude injection scenarios that are likely more representative of the height of the

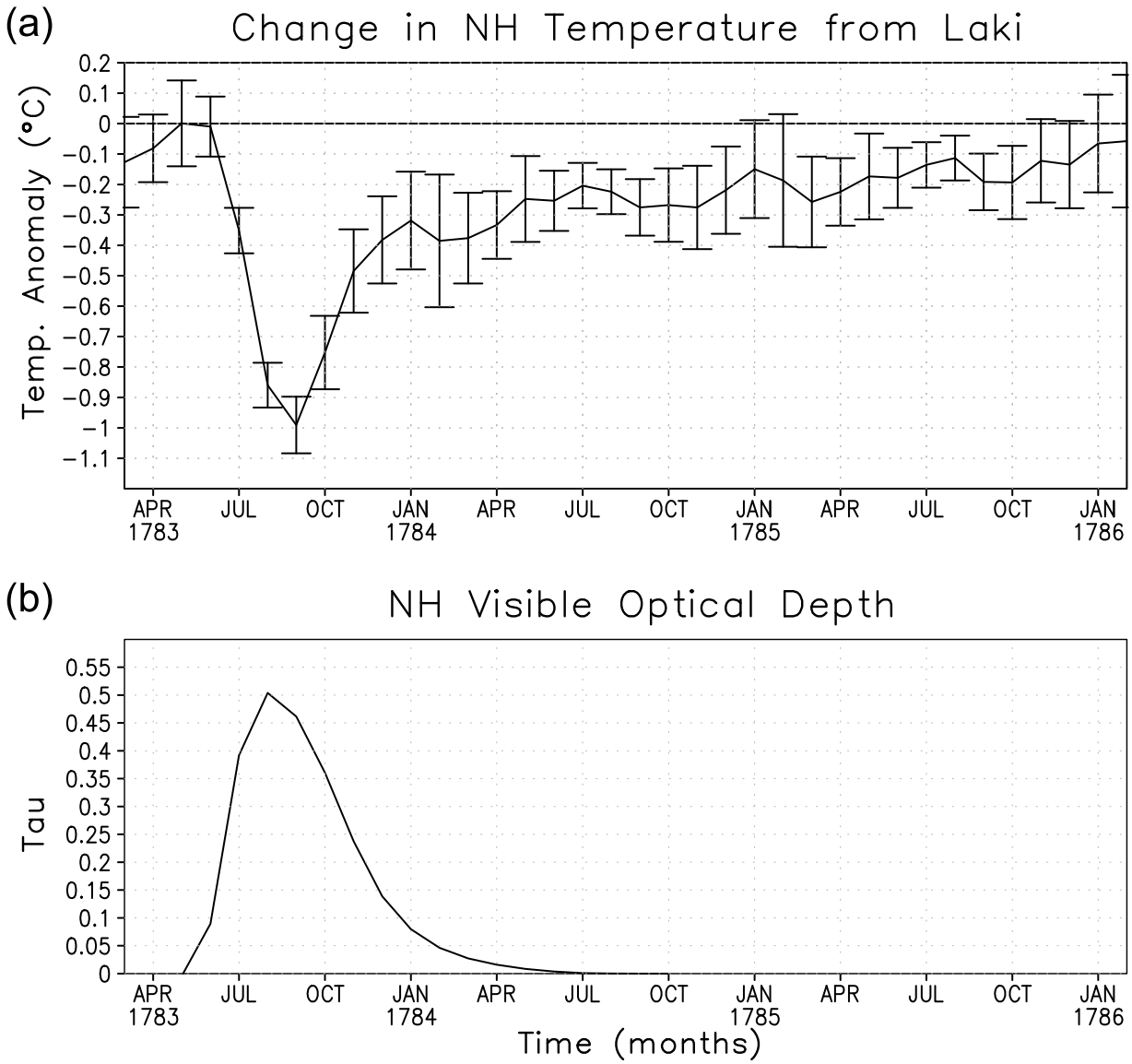


**Figure 4.** Simulated atmospheric and climatic perturbations due to the 1783–1784 CE Laki eruption with (a) showing the monthly mean temperature anomaly (°C) averaged over the Northern Hemisphere (NH) induced by Laki analyzing the model simulations of *Oman et al.* [2006a, 2006b]. The temperature anomaly is calculated with respect to the 30-year control run, and the error bars denote one standard deviation of the interannual variability from the control run; and (b) showing the sulfate optical depth averaged over the NH as a mean of the three Laki ensemble members used in *Oman et al.* [2006a].

eruption columns while Laki was most intensely active. Both *Oman et al.* [2006b] and *Highwood and Stevenson* [2003] concluded that their simulated summer temperature anomalies are in reasonable agreement with a variety of temperature proxy data [*Angell and Korshover*, 1985; *Briffa et al.*, 1998; *D'Arrigo and Jacoby*, 1999; *Jacoby et al.*, 1999] and the 1768–1794 European weather station records [*Manley*, 1974; *Kington*, 1988; *Parker et al.*, 1992; *Thordarson and Self*, 2003; *Brázdil et al.*, 2010].

[27] Overall, differences in surface temperature anomalies as evident from both the modeling studies and the proxy data imply that the surface temperature response during Laki was not spatially uniform. Therefore, it might not be straightforward to use proxy records of surface temperature changes for evaluating hemispheric mean temperature changes predicted by numerical models, with the response of regional surface temperatures to transient radiative effects depending on surface moisture, among other factors [see also *Mann et al.*, 2012 for additional discussion]. Furthermore, none of the previous Laki modeling studies fully quantified the aerosol indirect effect on climate following the eruption, which by analogy with present-day aerosol forcing of climate is expected to be at least of the same order of magnitude as the aerosol direct effect [e.g., *Lohmann and Feichter*, 2005; *Forster et al.*, 2007]. This is reinforced by the results of *Highwood and Stevenson* [2003] who estimated a NH mean

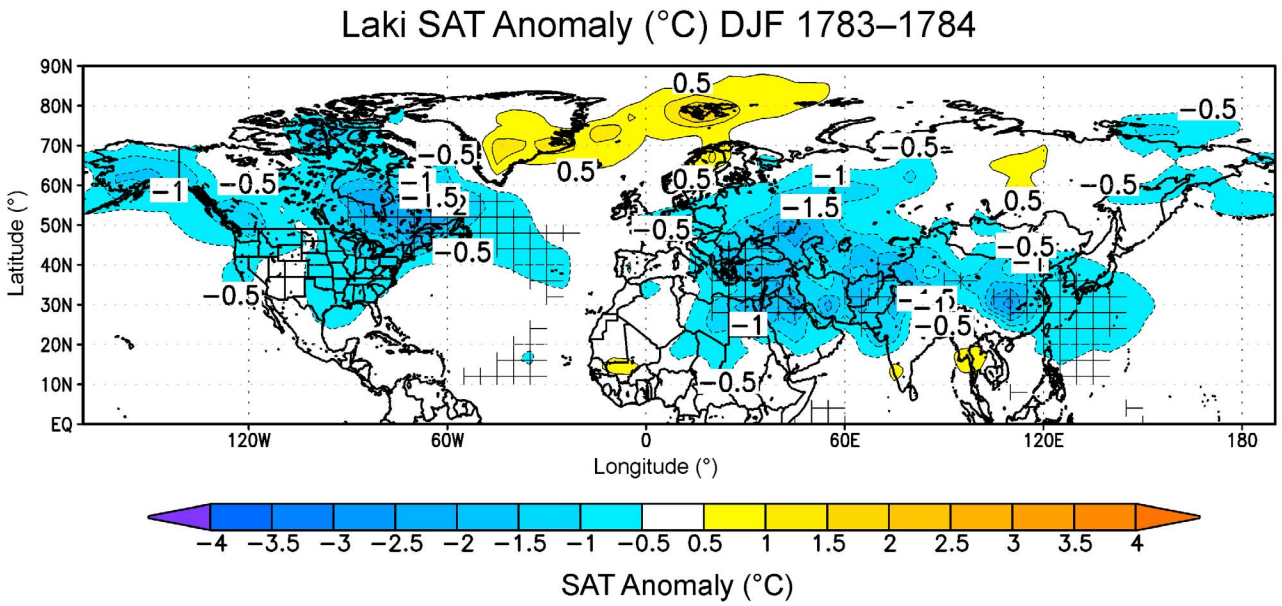


**Figure 5.** Surface air temperature (SAT) anomaly (°C) for the winter of 1783–1784 (DJF) induced by the Laki eruption using the model simulations by *Oman et al.* [2006a, 2006b]. The SAT anomaly is calculated with respect to the 30-year control run, and the hatching denotes the statistical significance at the 95% confidence level.

aerosol indirect effect of $-8$ W m$^{-2}$ during the first month of the eruption based on the purely tropospheric scenario.

[28] *D'Arrigo et al.* [2011] attributed the anomalously cold 1783–1784 winter to natural variability (rather than to the effects of the Laki aerosol), namely due to the occurrence of a negative North Atlantic Oscillation (NAO) combined with an El Niño event in the Pacific. This is a similar meteorological situation to that which occurred during the cold winter of 2009–2010. Using the GISS modelE, the heating rate anomalies induced by the Laki eruption did not produce a negative NAO [*Oman et al.*, 2006a, 2006b], nor do high-latitude eruptions in general appear to produce a negative Arctic Oscillation response in the model [*Oman et al.*, 2005]. Nonetheless, the direct radiative effects of the Laki aerosol produced statistically significant temperature anomalies during the winter of 1783–1784 in our simulations. Figure 5 shows calculated surface air temperature anomalies of up to $-0.5$°C over central Europe and up to $-2$°C over parts of eastern North America during the winter of 1783–1784 analyzing the model simulations by *Oman et al.* [2006a, 2006b]. Seasonally averaged over the NH we find a winter temperature anomaly of $-0.36$°C following Laki. For comparison, the 1991 eruption of Mt. Pinatubo resulted in a NH mean temperature anomaly of $-0.7$°C in the year 1992 [*McCormick et al.*, 1995, and references therein]. Therefore, all that can be deduced from the current evidence is that both the Laki radiative forcing during winter 1783–1784, albeit weaker than during the initial five months of the eruption, combined with the natural climate variability, caused the anomalous 1783–1784 winter temperatures. *Lanciki et al.* [2012] arrived at a similar conclusion, but for different reasons. Alternatively, the overall proposal by *Lanciki et al.* [2012] could be turned on its head by concluding that $\Delta^{33}$S measurements of the Laki $SO_4$ in one Greenland ice core are likely not valid evidence for any potential climate impact induced by a long-lasting high-latitude eruption in Iceland.

## 4. Conclusions and Future Work

[29] *Lanciki et al.* [2012] measured the sulfur isotopic composition of Laki sulfates deposited in a Greenland Summit ice core and found no significant $\Delta^{33}$S anomaly, which led them to conclude that Laki had "little or no stratospheric impact." We show that, based on previous work, the Laki sulfates would not be expected to possess a significant $\Delta^{33}$S anomaly, because the volatiles were not injected into stratospheric altitudes at which mass-independent fractionation of sulfur isotopes could be an important photochemical process (i.e., altitudes well above 17 km at high northern latitudes). *Lanciki et al.* [2012] used the lack of a $\Delta^{33}$S anomaly together with the apparently short aerosol deposition time interval (<6 months) to the Summit region to argue that the Laki eruption had "no significant climatic impact … beyond the initial effects ...." We suggest that the *Lanciki et al.* [2012] study does not categorically demonstrate that the Laki aerosol veil cannot have been present at upper tropospheric and lower stratospheric altitudes for more than six months because the majority of the Laki aerosol was transported by the polar jet stream to the east and northeast and eventually deposited at midlatitudes (Figures 1 and 3). Therefore, volcanic sulfate records in Greenland ice cores may be inadequate proxies for both the total sulfate deposition (hence the total atmospheric aerosol loading) and the duration of the presence of the aerosol veil in the upper troposphere for long-lasting, high-latitude eruptions.

[30] In conjunction with the simulated winter temperature anomalies following Laki (Figure 5) and temperature proxy data, one could conclude that $\Delta^{33}$S measurements of volcanic sulfates recorded in Greenland ice cores are not a valid

means to infer any potential climate impact a Laki-type flood lava event may have or have had. Such findings are important because $\Delta^{33}S$ measurements of sulfates formed after major explosive volcanic eruptions in the tropics injecting volatiles into altitudes >20 km may be used as an indicator for the climate relevance of an eruption [*Savarino et al.*, 2003; *Baroni et al.*, 2007; *Baroni et al.*, 2008; *Lanciki et al.*, 2012]. However, the applicability of sulfur isotopic measurements to interpret the climatic relevance of high-latitude eruptions in general has yet to be demonstrated. The 1912 Katmai eruption in Alaska might serve as a test case because it injected around 5 Tg of $SO_2$ into altitudes between 15 and 24 km [*Stothers*, 1996]. Even though the 1912 Katmai emissions reached altitudes at which photolysis of either $SO_2$ or $SO_3$ is expected, the volcanic aerosol cloud would need to be sufficiently spatially separated over time for a time-dependent anomaly to be detectable. A sufficient spatial separation is unlikely to be achieved for high-latitude eruptions due to the limited latitudinal dispersion of their volcanic aerosol clouds. One way forward might be to use photochemical models that account for isotopic fractionation [e.g., *Pavlov and Kasting*, 2002; *Pavlov et al.*, 2005] to simulate a range of high-latitude eruptions.

[31] One unresolved question about the Laki eruption is the very high temperatures experienced in western Europe during the summer of 1783 [e.g., *Luterbacher et al.*, 2004] that were contemporary with reports of a "dry fog" and the smell of sulfur coinciding with the Laki eruption [e.g., *Thordarson and Self*, 2003, and references therein]. Was this just a coincidence or were the warm temperatures somehow caused by the impacts of the Laki aerosol cloud? Climate model simulations did not reproduce the summer warming [*Oman et al.*, 2006a] nor does it seem that the radiative forcing from sulfur dioxide is large enough to explain the warm temperatures [*Highwood and Stevenson*, 2003]. Furthermore, the role of the warm temperatures in the summer of 1783 in inducing or contributing to the mortality crisis year of 1783 is not fully understood [e.g., *Durand and Grattan*, 1999; *Grattan et al.*, 2003; *Witham and Oppenheimer*, 2005; *Schmidt et al.*, 2011, and references therein]. Future work should also aim to make use of fully coupled chemistry-aerosol-climate models to investigate and reexamine potential feedbacks between, for example, the Laki aerosol loading in both the upper and lower troposphere and atmospheric circulation. There is also scope to further investigate whether Laki may or may not have contributed to the unusual atmospheric conditions and below-average winter temperatures that were observed not only in western Europe but also Russia and Japan during 1783–1784 [*Thordarson and Self*, 2003, and references therein], which cannot be explained solely by the occurrence of a negative NAO phase and an El Niño event.

[32] Governmental bodies in Europe are showing increasing interest in the development of hazard mitigation strategies in the event of another Laki-type eruption. Informed decisions can be made based on a probabilistic assessment of the impact of a Laki-type eruption on climate, atmospheric dynamics, human health, agriculture, ecosystems and commercial aviation. Such an assessment should aim to cover the full uncertainty range inherent in Laki-type volcanic events including, for example, the eruption column heights, the total amount of volatiles released, the influence of the atmospheric background state (both dynamic state and oxidant availability), the season of the eruption, and the role of volcanic ash in affecting chemical processes and the radiative forcing of climate. Thus it is important to examine all hypotheses bearing on the atmospheric impact of the 1783–1784 CE Laki eruption.

[33] **Acknowledgments.** We thank I.N. Bindeman and two other reviewers for their thoughtful comments and suggestions that helped to improve the manuscript. We also thank Sandip Dhomse for the provision of the SLIMCAT ozone data. Anja Schmidt is funded through NERC grant NE/I015612/1. Alan Robock is supported by NSF grants ATM-0730452 and AGS-1157525.

## References

Angell, J. K., and J. Korshover (1985), Surface temperature changes following the six major volcanic episodes between 1780 and 1980, *J. Clim. Appl. Meteorol.*, *24*(9), 937–951, doi:10.1175/1520-0450(1985)024<0937:STCFTS>2.0.CO;2.

Baroni, M., M. H. Thiemens, R. J. Delmas, and J. Savarino (2007), Mass-independent sulfur isotopic compositions in stratospheric volcanic eruptions, *Science*, *315*(5808), 84–87, doi:10.1126/science.1131754.

Baroni, M., J. Savarino, J. Cole-Dai, V. K. Rai, and M. H. Thiemens (2008), Anomalous sulfur isotope compositions of volcanic sulfate over the last millennium in Antarctic ice cores, *J. Geophys. Res.*, *113*(D20), D20112, doi:10.1029/2008JD010185.

Bindeman, I. N., J. M. Eiler, B. A. Wing, and J. Farquhar (2007), Rare sulfur and triple oxygen isotope geochemistry of volcanogenic sulfate aerosols, *Geochim. Cosmochim. Acta*, *71*(9), 2326–2343, doi:10.1016/j.gca.2007.01.026.

Brázdil, R., et al. (2010), European floods during the winter 1783/1784: Scenarios of an extreme event during the 'Little Ice Age,' *Theor. Appl. Climatol.*, *100*(1–2), 163–189, doi:10.1007/s00704-009-0170-5.

Briffa, K. R., P. D. Jones, F. H. Schweingruber, and T. J. Osborn (1998), Influence of volcanic eruptions on Northern Hemisphere summer temperature over the past 600 years, *Nature*, *393*(6684), 450–455, doi:10.1038/30943.

Burkholder, J. B., and S. McKeen (1997), UV absorption cross sections for $SO_3$, *Geophys. Res. Lett.*, *24*(24), 3201–3204, doi:10.1029/97GL03255.

Castleman, A. W., H. R. Munkelwitz, and B. Manowitz (1973), Contribution of volcanic sulphur compounds to the stratospheric aerosol layer, *Nature*, *244*(5415), 345–346, doi:10.1038/244345a0.

Castleman, A. W., H. R. Munkelwitz, and B. Manowitz (1974), Isotopic studies of the sulfur component of the stratospheric aerosol layer, *Tellus, Ser. A*, *26*(1–2), 222–234, doi:10.1111/j.2153-3490.1974.tb01970.x.

Chenet, A. L., F. Fluteau, and V. Courtillot (2005), Modelling massive sulphate aerosol pollution, following the large 1783 Laki basaltic eruption, *Earth Planet. Sci. Lett.*, *236*(3–4), 721–731, doi:10.1016/j.epsl.2005.04.046.

Chipperfield, M. P. (1999), Multiannual simulations with a three-dimensional chemical transport model, *J. Geophys. Res.*, *104*(D1), 1781–1805, doi:10.1029/98JD02597.

Chipperfield, M. P. (2006), New version of the TOMCAT/SLIMCAT off-line chemical transport model: Intercomparison of stratospheric tracer experiments, *Q. J. R. Meteorol. Soc.*, *132*, 1179–1203, doi:10.1256/qj.05.51.

Chung, K., J. G. Calvert, and J. W. Bottenheim (1975), The photochemistry of sulfur dioxide excited within its first allowed band (3130 Å) and the "forbidden" band (3700–4000 Å), *Int. J. Chem. Kinet.*, *7*(2), 161–182, doi:10.1002/kin.550070202.

Clausen, H. B., and C. U. Hammer (1988), The Laki and Tambora eruptions as revealed in Greenland ice cores from 11 locations, *Ann. Glaciol.*, *10*, 16–22.

Crowley, T. J., and M. B. Unterman (2012), Technical details concerning development of a 1200-yr proxy index for global volcanism, *Earth Syst. Sci. Data Discuss.*, *5*(1), 1–28, doi:10.5194/essdd-5-1-2012.

D'Arrigo, R. D., and G. C. Jacoby (1999), Northern North American tree-ring evidence for regional temperature changes after major volcanic events, *Clim. Change*, *41*, 1–15, doi:10.1023/A:1005370210796.

D'Arrigo, R., R. Seager, J. E. Smerdon, A. N. LeGrande, and E. R. Cook (2011), The anomalous winter of 1783–1784: Was the Laki eruption or an analog of the 2009–2010 winter to blame?, *Geophys. Res. Lett.*, *38*, L05706, doi:10.1029/2011GL046696.

Durand, M., and J. Grattan (1999), Extensive respiratory health effects of volcanogenic dry fog in 1783 inferred from European documentary sources, *Environ. Geochem. Health*, *21*, 371–376, doi:10.1023/A:1006700921208.

Farquhar, J., J. Savarino, S. Airieau, and M. H. Thiemens (2001), Observation of wavelength-sensitive mass-independent sulfur isotope effects during $SO_2$ photolysis: Implications for the early atmosphere, *J. Geophys. Res.*, *106*(E12), 32,829–32,839, doi:10.1029/2000JE001437.

Forster, P. et al. (2007), Changes in atmospheric constituents and in radiative forcing, in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by S. Solomon et al., pp. 129–234, Cambridge Univ. Press, Cambridge, U. K.

Gao, C., L. Oman, A. Robock, and G. L. Stenchikov (2007), Atmospheric volcanic loading derived from bipolar ice cores: Accounting for the spatial distribution of volcanic deposition, *J. Geophys. Res.*, *112*, D09109, doi:10.1029/2006JD007461.

Gao, C., A. Robock, and C. Ammann (2008), Volcanic forcing of climate over the past 1500 years: An improved ice core-based index for climate models, *J. Geophys. Res.*, *113*, D23111, doi:10.1029/2008JD010239.

Grattan, J., M. Durand, and S. Taylor (2003), Illness and elevated human mortality in Europe coincident with the Laki Fissure eruption, in *Volcanic Degassing*, edited by C. Oppenheimer, D. M. Pyle, and J. Barclay, *Geol. Soc. Spec. Publ.*, *213*, 401–414.

Highwood, E. J., and D. S. Stevenson (2003), Atmospheric impact of the 1783–1784 Laki eruption: Part II—Climatic effect of sulphate aerosol, *Atmos. Chem. Phys.*, *3*, 1177–1189, doi:10.5194/acp-3-1177-2003.

Jacoby, G. C., K. W. Workman, and R. D. D'Arrigo (1999), Laki eruption of 1783, tree rings, and disaster for northwest Alaska Inuit, *Quat. Sci. Rev.*, *18*(12), 1365–1371, doi:10.1016/S0277-3791(98)00112-7.

Kington, J. A. (1988), *The Weather of the 1780's Over Europe*, Cambridge Univ. Press, New York, doi:10.1017/CBO9780511735721.

Koch, D., G. A. Schmidt, and C. V. Field (2006), Sulfur, sea salt and radionuclide aerosols in GISS ModelE, *J. Geophys. Res.*, *111*, D06206, doi:10.1029/2004JD005550.

Kravitz, B., and A. Robock (2011), Climate effects of high-latitude volcanic eruptions: Role of the time of year, *J. Geophys. Res.*, *116*, D01105, doi:10.1029/2010JD014448.

Lacasse, C. (2001), Influence of climate variability on the atmospheric transport of Icelandic tephra in the subpolar North Atlantic, *Global Planet. Change*, *29*(1–2), 31–55, doi:10.1016/S0921-8181(01)00099-6.

Lanciki, A., J. Cole-Dai, M. H. Thiemens, and J. Savarino (2012), Sulfur isotope evidence of little or no stratospheric impact by the 1783 Laki volcanic eruption, *Geophys. Res. Lett.*, *39*, L01806, doi:10.1029/2011GL050075.

Lohmann, U., and J. Feichter (2005), Global indirect aerosol effects: A review, *Atmos. Chem. Phys.*, *5*(3), 715–737, doi:10.5194/acp-5-715-2005.

Luterbacher, J., D. Dietrich, E. Xoplaki, M. Grosjean, and H. Wanner (2004), European seasonal and annual temperature variability, trends, and extremes since 1500, *Science*, *303*(5663), 1499–1503, doi:10.1126/science.1093877.

Manley, G. (1974), Central England temperatures: Monthly means 1659 to 1973, *Q. J. R. Meteorol. Soc.*, *100*(425), 389–405, doi:10.1002/qj.49710042511.

Mann, M. E., J. D. Fuentes, and S. Rutherford (2012), Underestimation of volcanic cooling in tree-ring-based reconstructions of hemispheric temperatures, *Nat. Geosci.*, *5*, 202–205, doi:10.1038/ngeo1394.

Martin, E., and I. Bindeman (2009), Mass-independent isotopic signatures of volcanic sulfate from three supereruption ash deposits in Lake Tecopa, California, *Earth Planet. Sci. Lett.*, *282*(1–4), 102–114, doi:10.1016/j.epsl.2009.03.005.

McCormick, M. P., L. W. Thomason, and C. R. Trepte (1995), Atmospheric effects of the Mt Pinatubo eruption, *Nature*, *373*(6513), 399–404, doi:10.1038/373399a0.

Mosley-Thompson, E., T. A. Mashiotta, and L. G. Thompson (2003), High-resolution ice core records of late Holocene volcanism: Current and future contributions from the Greenland PARCA cores, in *Volcanism and the Earth's Atmosphere*, *Geophys. Monogr. Ser.*, vol. 139, edited by A. Robock and C. Oppenheimer, pp. 153–164, AGU, Washington, D. C.

Oman, L., A. Robock, G. Stenchikov, G. A. Schmidt, and R. Ruedy (2005), Climatic response to high latitude volcanic eruptions, *J. Geophys. Res.*, *110*, D13103, doi:10.1029/2004JD005487.

Oman, L., A. Robock, G. L. Stenchikov, T. Thordarson, D. Koch, D. T. Shindell, and C. Gao (2006a), Modeling the distribution of the volcanic aerosol cloud from the 1783–1784 Laki eruption, *J. Geophys. Res.*, *111*, D12209, doi:10.1029/2005JD006899.

Oman, L., A. Robock, G. L. Stenchikov, and T. Thordarson (2006b), High-latitude eruptions cast shadow over the African monsoon and the flow of the Nile, *Geophys. Res. Lett.*, *33*, L18711, doi:10.1029/2006GL027665.

Oppenheimer, C. (2003), Ice core and palaeoclimatic evidence for the timing and nature of the great mid-13th century volcanic eruption, *Int. J. Climatol.*, *23*(4), 417–426, doi:10.1002/joc.891.

Parker, D. E., T. P. Legg, and C. K. Folland (1992), A new daily central England temperature series, 1772–1991, *Int. J. Climatol.*, *12*(4), 317–342, doi:10.1002/joc.3370120402.

Pavlov, A. A., and J. F. Kasting (2002), Mass-Independent fractionation of sulfur isotopes in Archean sediments: Strong evidence for an anoxic Archean atmosphere, *Astrobiology*, *2*(1), 27–41, doi:10.1089/153110702753621321.

Pavlov, A. A., M. J. Mills, and O. B. Toon (2005), Mystery of the volcanic mass-independent sulfur isotope fractionation signature in the Antarctic ice core, *Geophys. Res. Lett.*, *32*, L12816, doi:10.1029/2005GL022784.

Petersen, G. N., H. Bjornsson, and P. Arason (2012), The impact of the atmosphere on the Eyjafjallajökull 2010 eruption plume, *J. Geophys. Res.*, *117*, D00U07, doi:10.1029/2011JD016762.

Robock, A. (2000), Volcanic eruptions and climate, *Rev. Geophys.*, *38*(2), 191–219, doi:10.1029/1998RG000054.

Savarino, J., A. Romero, J. Cole-Dai, S. Bekki, and M. H. Thiemens (2003), UV induced mass-independent sulfur isotope fractionation in stratospheric volcanic sulfate, *Geophys. Res. Lett.*, *30*(21), 2131, doi:10.1029/2003GL018134.

Schmidt, G. A., et al. (2006), Present-day atmospheric simulations using GISS ModelE: Comparison to in situ, satellite, and reanalysis data, *J. Clim.*, *19*(2), 153–192, doi:10.1175/JCLI3612.1.

Schmidt, A., K. S. Carslaw, G. W. Mann, M. Wilson, T. J. Breider, S. J. Pickering, and T. Thordarson (2010), The impact of the 1783–1784 AD Laki eruption on global aerosol formation processes and cloud condensation nuclei, *Atmos. Chem. Phys.*, *10*(13), 6025–6041, doi:10.5194/acp-10-6025-2010.

Schmidt, A., B. Ostro, K. S. Carslaw, M. Wilson, T. Thordarson, G. W. Mann, and A. J. Simmons (2011), Excess mortality in Europe following a future Laki-style Icelandic eruption, *Proc. Natl. Acad. Sci. U. S. A.*, *108*(38), 15,710–15,715, doi:10.1073/pnas.1108569108.

Self, S., and M. Rampino (2012), The 1963–1964 eruption of Agung volcano (Bali, Indonesia), *Bull. Volcanol.*, *74*(6), 1521–1536, doi:10.1007/s00445-012-0615-z.

Self, S., R. Gertisser, T. Thordarson, M. R. Rampino, and J. A. Wolff (2004), Magma volume, volatile emissions, and stratospheric aerosols from the 1815 eruption of Tambora, *Geophys. Res. Lett.*, *31*, L20608, doi:10.1029/2004GL020925.

Stephensen, O. (1783), Abstract from Prefect Stephensen letter to Erichsen, the deputy of the treasury, in *Skaftareldar 1783–84: Ritgerdir og Heimildir*, p. 279, Mál og Menning, Reykjavik.

Stevenson, D. S., C. E. Johnson, E. J. Highwood, V. Gauci, W. J. Collins, and R. G. Derwent (2003), Atmospheric impact of the 1783–1784 Laki eruption: Part I Chemistry modelling, *Atmos. Chem. Phys.*, *3*, 487–507, doi:10.5194/acp-3-487-2003.

Stothers, R. B. (1984), The Great Tambora eruption in 1815 and its aftermath, *Science*, *224*(4654), 1191–1198, doi:10.1126/science.224.4654.1191.

Stothers, R. B. (1996), Major optical depth perturbations to the stratosphere from volcanic eruptions: Pyrheliometric period, 1881–1960, *J. Geophys. Res.*, *101*(D2), 3901–3920, doi:10.1029/95JD03237.

Stothers, R. B., J. A. Wolff, S. Self, and M. R. Rampino (1986), Basaltic fissure eruptions, plume heights, and atmospheric aerosols, *Geophys. Res. Lett.*, *13*(8), 725–728, doi:10.1029/GL013i008p00725.

Thordarson, T., and G. Larsen (2007), Volcanism in Iceland in historical time: Volcano types, eruption styles and eruptive history, *J. Geodyn.*, *43*(1), 118–152, doi:10.1016/j.jog.2006.09.005.

Thordarson, T., and S. Self (1993), The Laki (Skaftar Fires) and Grimsvötn eruptions in 1783–1785, *Bull. Volcanol.*, *55*(4), 233–263, doi:10.1007/BF00624353.

Thordarson, T., and S. Self (2003), Atmospheric and environmental effects of the 1783–1784 Laki eruption: A review and reassessment, *J. Geophys. Res.*, *108*(D1), 4011, doi:10.1029/2001JD002042.

Thordarson, T., S. Self, N. Oskarsson, and T. Hulsebosch (1996), Sulfur, chlorine, and fluorine degassing and atmospheric loading by the 1783–1784 AD Laki (Skaftar fires) eruption in Iceland, *Bull. Volcanol.*, *58*(2–3), 205–225, doi:10.1007/s004450050136.

Thordarson, T., D. J. Miller, G. Larsen, S. Self, and H. Sigurdsson (2001), New estimates of sulfur degassing and atmospheric mass-loading by the 934 AD Eldgja eruption, Iceland, *J. Volcanol. Geothermal Res.*, *108*(1–4), 33–54, doi:10.1016/S0377-0273(00)00277-8.

Thordarson, T., S. Self, D. J. Miller, G. Larsen, and E. G. Vilmundardottir (2003), Sulphur release from flood lava eruptions in the Veidivotn, Grimsvotn and Katla volcanic systems, Iceland, in *Volcanic Degassing*, edited by C. Oppenheimer, D. M. Pyle, and J. Barclay, *Geol. Soc. Spec. Publ.*, *213*, 103–121.

Thordarson, T., M. Rampino, L. P. Keszthelyi, and S. Self (2009), *Effects of Megascale Eruptions on Earth and Mars*, *Spec. Pap. Geol. Soc. Am.*, *453*, 37–53, doi:10.1130/2009.453(04).

UK National Risk Register (2012), *National Risk Register of Civil Emergencies*, 55 pp., Cabinet Off., London.

Witham, C. S., and C. Oppenheimer (2005), Mortality in England during the 1783–4 Laki Craters eruption, *Bull. Volcanol.*, *67*(1), 15–26, doi:10.1007/s00445-004-0357-7.

Woods, A. W. (1993), A model of the plumes above basaltic fissure eruptions, *Geophys. Res. Lett.*, *20*, 1115–1118, doi:10.1029/93GL01215.